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ADJUSTMENT TO A MOIST ADIABATIC LAPSE IN THE PEP MODEL

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Whenever convective overturnings occur in the atmosphere, whether in isolated thunderstorms or imbedded within larger scale phenomena, there will be a vertical exchange of momentum and thermal energy. To a first crude but simple approximation, we may say that the resulting wind and temperature structure will be one of no shear in the wind components and a moist adiabatic lapse of temperature.

This crude approximation is modeled in the PEP by taking advantage of the three layers of precipitation information carried, the boundary layer and the two lower troposphere layers. The procedure is an iterative one and starts at the lowest layer. If the layer is saturated, the pseudo equivalent potential temperature θ_{SE} defined by that layer's temperature and pressure is computed. (If the layer is not saturated, nothing further is done and consideration passes up to the next layer.) Assuming the layer to be saturated, the θ_{SE} is used to compute the temperature a rising parcel would have if lifted along that pseudo-adiabat (θ_{SE}) to the middle of the layer above. The method developed by Stackpole (JAM V6 #3 p.464) is employed for this.

The lifted parcel temperature is then compared with the temperature forecast for the layer into which it was lifted. If the lifted temperature is less than the layer temperature, i.e. the two layers are stable, nothing further is done and again consideration passes up to the next layer and the processes repeat.

If, however, the lifted temperature is warmer than the layer, i.e. the lower layer is convectively unstable with respect to the one above it, adjustments to the temperature and winds in the layers takes place. If θ_U and $p_{\sigma U}$ represent the potential temperature and pressure thickness of the upper layer, θ_L and $p_{\sigma L}$ those of the lower layer and $\delta\theta$ the temperature difference (θ_U - lifted temperature) (a negative quantity for the unstable conditions we are considering) then the adjusted temperature in the upper layer is given by

$$\theta_{AU} = \theta_U - \delta\theta \frac{p_{\sigma L}}{p_{\sigma U} + p_{\sigma L}}$$

and the lower

$$\theta_{AL} = \theta_L + \delta\theta \frac{p_{\sigma U}}{p_{\sigma U} + p_{\sigma L}}$$

What this amounts to is that the lower layer is cooled and the upper warmed by an amount weighted inversely with the amount of mass (p_σ representing the mass of the layer) involved. For example, if the lower layer were the boundary layer ($p_{\sigma L} = 50$ mb and $p_{\sigma U} \doteq 200$ mb) most of the temperature change would take place in the thinner lower layer. On the other hand, two equal layers would each participate equally in the temperature adjustment.

The wind adjustment is somewhat simpler in line with the assumption that the result will be zero shear. The wind components in each layer are equated to an adjusted wind given by

$$\mathbb{V}_A = \frac{\mathbb{V}_L \, p_{\sigma L} + \mathbb{V}_U \, p_{\sigma U}}{p_{\sigma L} + p_{\sigma U}}$$

i.e. just the mass weighted average of the winds in the lower ($\mathbb{V}_L$) and upper ($\mathbb{V}_U$) layers respectively.